Some Chemical Features of Lavas from the Manu'a Islands, Samoa[1]

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ABSTRACT: The lavas of the Manu'a Islands have chemical compositions typical of oceanic island alkali lavas. They have rare earth abundances with chondrite normalized lanthanum:ytterbium ratios around 10 and ytterbium concentrations about 2.2 ppm. Strontium concentrations are greater than 400 ppm and within the range of strontium values for Hawaiian alkalic lavas. Potassium:rubidium ratios are generally between 300 and 400 but three samples have potassium:rubidium $\sim$ 700, suggesting heterogeneity of source materials. The ^{87}strontium:^{86}strontium ratios average 0.7046 $\pm$ .0003, and are the highest known for oceanic islands. Low pressure differentiation is controlled by olivine and plagioclase. These lavas were segregated from a normal oceanic upper mantle at $>$40 km depth and the percentage of partial melting was in the range of 3 to 7 percent. The chemical composition of the probable original magma is estimated using combined trace element, major element, and partition coefficient data.

THE SAMOAN ISLANDS are located at the northern end of a major geological feature in the South Pacific, the Tonga-Kermadec Trench. The Tonga-Kermadec Trench has been interpreted as the site of large scale underthrusting (Isacks, Oliver, and Sykes, 1968), whereas the Samoan Islands are "normal" oceanic islands similar in tectonic setting to the Hawaiian Islands (Stice, 1968; Macdonald, 1944). In view of the differences in tectonic setting and the geographic proximity of these two areas, it is important that the chemical features of Samoan lavas be well defined so that comparisons with Tongan lavas can be confidently made. In addition, the Samoan Islands are chemically interesting of their own accord for the Manu'a Islands exhibit chemical variations that can be deciphered by the combined use of major element, trace element, and partition coefficient data.

Stice (1968) described the geology and petrology of the Manu'a Islands, Samoa. Samples donated by Stice have been used in this study. Twenty samples were analyzed for zirconium, nickel, strontium, and rubidium by X-ray fluorescence (XRF) and five were analyzed for rare earth elements and potassium:rubidium by stable isotope dilution mass spectrometry: ^{87}Sr:^{86}Sr ratios were also measured on these samples. Use of these trace element and isotopic data in conjunction with already published major element data (Stice, 1968) allows these lavas to be characterized and understood more thoroughly than is possible by major element or trace element data alone. The results of this characterization will be interpreted according to current geochemical models and ideas (Gast, 1968; Hubbard, 1969; Hubbard, Gast, and Sun, in prep.; Tilley, Yoder, and Schairer, 1964–1965), and contrasted with some of the results of Hubbard and Gast (in prep.) for the Tonga Islands.

PRESENTATION OF DATA

Rare Earth Elements

The rare earth element (REE) abundance patterns (Fig. 1) are the light rare earth enriched patterns with low ytterbium concentrations characteristic of all analyzed oceanic island alkalic basalts (Hubbard, 1969; Frey et al., 1968). The chondrite normalized lanthanum:ytterbium ratios range from about 7.5 to about

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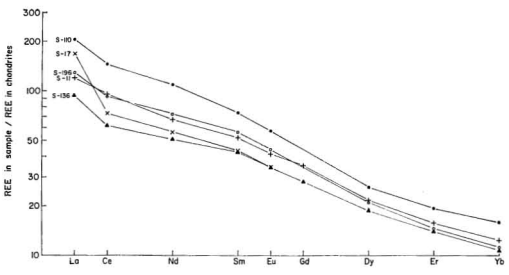

FIG. 1. Rare earth abundance patterns for five Samoan lavas. The individual REE values have been divided by the chondrite REE values of Frey et al. (1968).

12.8 and increase with increasing total rare earth concentrations. No europium anomaly is found, not even in a feldspar-phyric basalt (S-110) which has 45 percent plagioclase phenocrysts (Stice, 1968). These light REE-enriched abundance patterns are grossly different from the REE abundance patterns of lavas from the Tonga Islands (Hubbard and Gast, in prep.) which have patterns almost identical with oceanic ridge basalts (Kay, Hubbard, and Gast, 1970; Frey et al., 1968; Hubbard, 1969).

Strontium Concentrations and ^{87}Strontium:^{86}Strontium Ratios

The $^{87}Sr:^{86}Sr$ ratio is 0.7046 ± 0.0003 for five samples (Table 1). This agrees very well with the average ratio of 0.7046 found by Swainbank (1967) for four other Samoan samples. These are the highest $^{87}Sr:^{86}Sr$ ratios presently known for any oceanic island basalts, tholeiitic or alkalic, and are measurably higher than any presently measured for Tongan lavas (Hubbard and Gast, in prep.).

Total Sr concentrations range from 400 to 550 ppm in the basalts and up to 680 ppm in the hawaiites. One unusual feldspar-rich (45 percent) sample has 960 ppm Sr. The Sr concentrations are similar to those found in Hawaiian alkalic lavas, where alkalic basalts have 400 to 800 ppm Sr and hawaiites up to 1,300 ppm Sr (Hubbard, 1967). By contrast Tongan lavas have less than 290 ppm Sr, even in differentiated lavas (Hubbard and Gast, in prep.).

Strontium is plotted against weight percent CaO in Figure 2, against P_2O_5 in Figure 3, and against Al_2O_3 in Figure 4. It is obvious that Sr bears no correlation to Ca in the whole rock, although there is a fair correlation with Al_2O_3 and, therefore, perhaps with feldspar. The relatively good covariation of Sr and P is identical to the situation in Hawaiian alkalic lavas (Hubbard, 1967).

The one feldspar-rich sample, S-110, shows more Sr than expected from its P_2O_5 concentration (see Henderson, 1968 and Anderson and Greenland, 1969 for the use of P_2O_5 as an index of differentiation). The P_2O_5 value reported by Stice (1968) is 69 percent of the P_2O_5 value that corresponds to the 960 ppm Sr found for S-110, assuming the linear rela-

TABLE 1

RARE EARTH ELEMENTS, $^{87}Sr:^{86}Sr$, AND ISOTOPE DILUTION K:Rb DATA

Element	Precision	S-11	S-17	S-110	S-136	S-196
K(%)	±4.0%	0.71	0.75	1.58	0.57	0.71
Rb(ppm)	±4.0%	10.2	20.1	42.6	8.1	10.9
K:Rb		696	373	371	704	651
$^{87}Sr:^{86}Sr$*	±.0005	—	0.7048	0.7044	0.7041	0.7045
La(ppm)	±5.0%	39.5	55.9**	69.7	27.5	42.6
Ce(ppm)	±5.0%	75.4	64.5	127.2	54.1	82.4
Nd(ppm)	±5.0%	40.2	33.8	71.2	30.6	43.6
Sm(ppm)	±5.0%	9.46	7.90	13.4	7.75	10.2
Eu(ppm)	±5.0%	2.86	2.37	3.96	2.37	3.04
Gd(ppm)	±10.0%	8.82	—	—	7.04	—
Dy(ppm)	±5.0%	7.38	—	8.70	6.26	7.10
Er(ppm)	±5.0%	3.17	—	3.87	2.78	2.94
Yb(ppm)	±5.0%	2.49	2.15	3.18	2.16	2.26

* S-97:0.7051.

** This lanthanum value may be too high; see Figure 1.

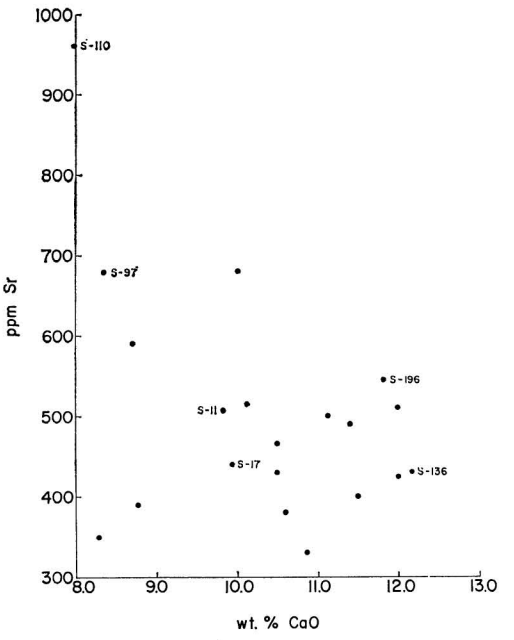

FIG. 2. Sr vs. CaO for 20 Samoan lavas. The six samples for which REE, K:Rb, and $^{87}Sr:^{86}Sr$ analyses were made are noted by their sample numbers.

tionship of Sr and P_2O_5 found for the other Manu'a samples (Fig. 3). This could be explained as the reduction of the P_2O_5 concentration by dilution with ~31 volume percent of

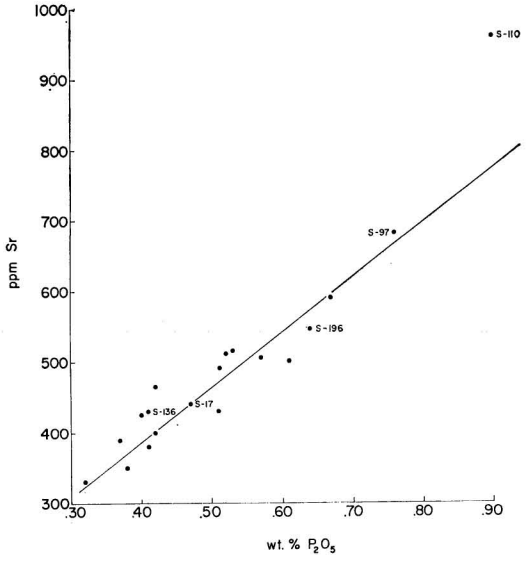

FIG. 3. Sr vs. P_2O_5 for the samples of Figure 2.

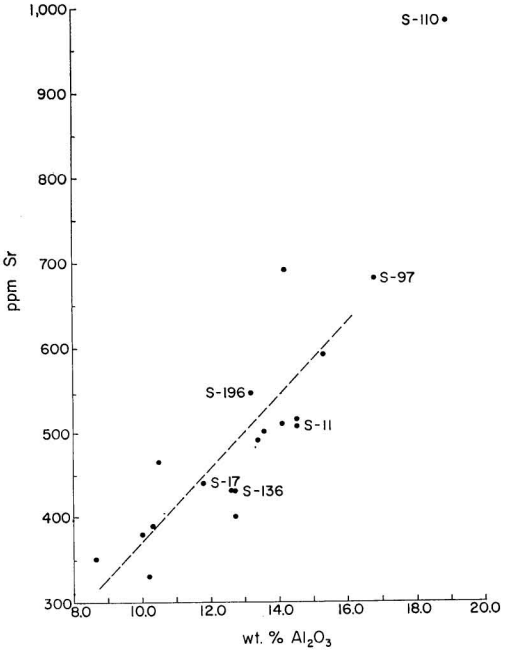

FIG. 4. Sr vs. Al_2O_3 for the samples of Figure 2.

plagioclase. However, there is another explanation for which a mass balance calculation can be made using the distribution coefficient of 1.5 for Sr plagio/Sr liquid found by Philpotts and Schnetzler (1970), the percentage plagioclase phenocrysts found by Stice, and the bulk Sr concentration (960 ppm) found by this study. This calculation shows that the plagioclase has 1,170 ppm Sr and the rest of the rock ("liquid") has 780 ppm Sr. This calculation is valid only if the phenocrysts crystallized *in situ* and did not result from addition of plagioclase phenocrysts. If the 45 percent plagioclase phenocrysts were the result of adding plagioclase crystals to the magma, then the data of Schnetzler and Philpotts (1968) show that there should be a +13 percent Eu anomaly in the REE data. The maximum Eu anomaly is ~3 to 4 percent (within the limits of error) indicating that no more than ~10 percent of plagioclase was added to the magma. Thus the bulk of the plagioclase phenocrysts crystallized *in situ* from a high Al_2O_3 magma. The discrepancy between this conclusion and the indication from the P_2O_5 data of plagioclase addition can be explained by noting that S-110 and S-97 (a

hawaiite) are essentially the same in their bulk chemical compositions and amount to total model feldspar—69 and 67 percent, respectively (Stice, 1968), although S-110 has 45 percent plagioclase phenocrysts. Therefore, the basic difference between S-97 and S-110 is that S-110 was subjected to conditions under which extensive plagioclase crystallized as phenocrysts. Under such conditions the P_2O_5 index may have been negated by crystallization and removal of low Sr apatite. In this case the linear Sr:P_2O_5 covariation found for the other Manu'a samples should not apply to S-110.

Figure 5 shows weight percent Al_2O_3 vs. weight percent MgO. The most MgO-rich samples are picrites or picritic (Stice, 1968). The most Al_2O_3-rich sample, S-110, is feldspar-rich, and S-97 is a hawaiite. This control of the bulk chemical composition of the alkalic lavas by the crystallization and movement of olivine and plagioclase is consistent with the 1-atmosphere melting experiments of Tilley, Yoder, and Schairer (1964–1965) which show that the crystallization sequence with decreasing temperature for alkalic basalts is olivine first and plagioclase second. Thus, the covariation of Sr and Al_2O_3 seen in Figure 4 is partly due to the enrichment of Al_2O_3 and Sr in the liquids derived from low pressure crystallization and removal of olivine, and partly due to the dilution of Al_2O_3 and Sr by the addition of olivine.

Nickel and Magnesium Oxide Concentrations

No discussion of Ni in basaltic lavas is very meaningful without relating the Ni concentrations to MgO concentrations. Figure 6 shows Ni vs. weight percent MgO on a linear graph. The data show uncommonly good correlation between Ni and MgO. The outstanding features of this data plot are: (1) the data points deviate from a straight line at MgO concentrations near those of the intersection of the olivine and plagioclase control lines of Figure 5; and (2) data points for Ni greater than 300 ppm do not fall along a single straight line.

Olivines separated from two picrites and analyzed by XRF have 2,920 and 3,120 ppm Ni and 14.8 weight percent FeO. Taking 3,000 ppm Ni for the olivines, which Stice (1968)

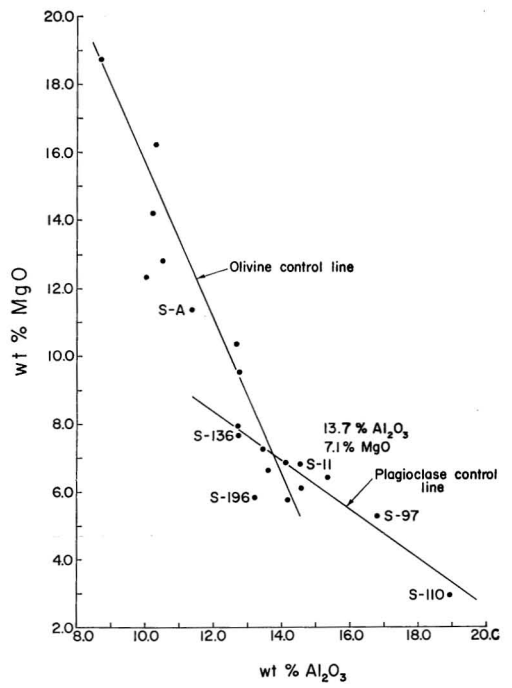

FIG. 5. Al_2O_3 vs. MgO for the samples of Figure 2. The olivine and plagioclase control lines are shown as are the MgO and Al_2O_3 values at their intersection.

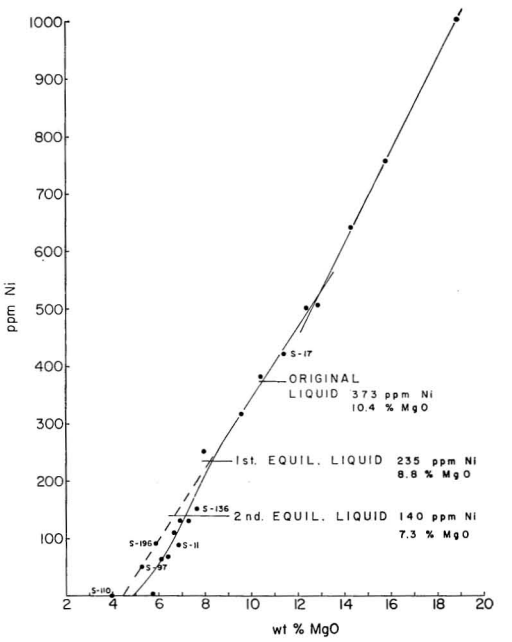

FIG. 6. Ni vs. MgO for the samples of Figure 2. The original liquid, 1st equilibration liquid, and 2nd equilibration liquid are shown.

describes as homogeneous in the picrites, and the distribution coefficient of 12.8 for Ni olivine:Ni liquid, derived from the data of Häkli and Wright (1967) by Hubbard, Gast, and Sun (in prep.) for 1,200° C, the liquid (first equilibration liquid) with which these olivines equilibrated had 235 ppm Ni. If the liquid from which these olivines crystallized was capable of crystallizing 5 volume percent of olivine, then the Ni concentration of the liquid before the onset of olivine crystallization was 373 ppm Ni. Less olivine will, of course, give lower Ni concentrations. These calculations apply only to the five most Ni-rich samples which lie along the upper straight line in Figure 6. Figure 7 shows Ni and MgO data for the five most Ni-rich samples and for the two olivines. These six points form a straight line and show clearly that, for Ni greater than 470 ppm, the Ni and MgO variations are the result of mixing these olivines and a liquid. The liquid was calculated to have 235 ppm Ni and it may be seen from this graph that this liquid would be 8.8 percent MgO. This MgO concentration is considerably greater than that of the intersection of the olivine and plagioclase control lines of Figure 5 and, thus, this intersection cannot represent the primary magma.

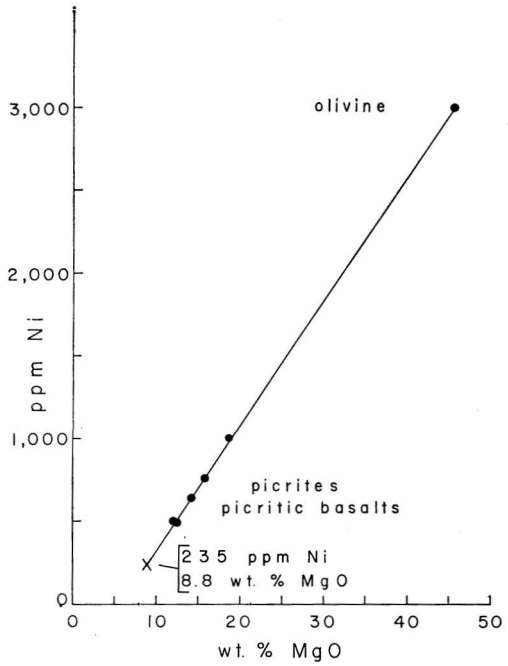

FIG. 7. Ni vs. MgO for the five most Ni-rich samples and the two olivines. The 1st equilibration liquid is noted.

In Figure 6 it is shown that, when Ni is equal to 300 to 470 ppm, the data can be fitted with a straight line having a lesser slope than the line which passes through the data which apply when nickel is more than 470 ppm and for the olivine data. This indicates a second episode of olivine crystallization in which the olivine was less Ni- and/or MgO-rich. By analogy with data for Kilauea Volcano (Murata and Richter, 1966; Richter et al., 1964; Moore and Krivoy, 1964), if the first episode of olivine crystallization occurred at about 1,200° C, the second would occur at about 1,150° C. In any case, the second-generation olivines would have crystallized at lower temperatures and from a liquid that contained only 235 ppm Ni. Thus, these olivines will be less Ni- and MgO-rich. The distribution coefficient for Ni olivine:Ni liquid increases with decreasing temperature from 12.8 at 1,200° C to about 14.0 at 1,150° C. Assuming that only 2 volume percent of olivine crystallized, it can be calculated that the second generation olivine will have 1,900 ppm Ni and its equilibrium liquid will have 140 ppm Ni. This 140 ppm Ni corresponds to about 7.3 percent MgO, which agrees well with the 7.1 percent MgO that corresponds to the intersection of the olivine-plagioclase control lines of Figure 5.

If the original liquid was able to crystallize 5 volume percent olivine, then its MgO concentration was near 10.4 percent. Greater MgO concentrations, therefore, must be due to the addition of olivine to the original liquid. This can be done only by removing olivine from one portion of the magma or lava volume and adding it to a second, smaller volume. Stice (1968) reported that some picrites contain over 65 percent olivine. This would correspond to removing the assumed 5 volume percent of olivine from about 93 percent of the magma volume and adding it to the remaining 7 percent.

Stice (1968) noted that the olivines in the dunite nodules found in some Samoan lavas have the same composition as the olivines in

the picrites. He interpreted these nodules to be concentrations of olivine crystals that resulted from olivine settling out of the magma (see Figure 7 for supporting evidence presented by this report). Murata and Richter (1966) interpreted observed correlations of eruption rate and olivine content for the 1959 Kilauea eruption as being due to the erosion of olivine accumulations in the vent system and magna chamber. Clearly, to erupt lavas with high contents of olivine crystals, flow rates in the erupting lava must be great enough to keep the crystals suspended in the liquid. Kilauea Volcano, in common with other Hawaiian volcanoes, typically erupts large volumes of lava in a short time from a fairly localized vent or portion of a rift system (Eaton and Murata, 1960; Moore and Krivoy, 1964; Richter et al., 1964; Macdonald and Eaton, 1964). It should not, therefore, be surprising that flow rates within the erupting magma body can be relatively high and that transport of solid material (olivine crystals) may be quite pronounced. The Manu'a lavas were erupted from a vent and rift system similar to that found at Kilauea; therefore, similarly rapid rates of eruption are probable on this basis alone. Note also that Hawaiian alkalic basalts have often transported ultramafic nodules several inches in diameter in cross section to the surface (White, 1966; Jackson, 1966), a feat which requires even greater flow rates than does the transport of olivine crystals. Greater magma viscosities could also provide greater transporting ability, but this effect is absent or small as indicated by the predominance of thin lava flows on the Manu'a Islands (Macdonald, 1944; Stice, 1968). These MgO- and olivine-rich lavas are, therefore, cumulates and do not provide evidence for picritic magmas.

Potassium:Rubidium Ratios

One of the unusual features of this collection of Samoan lavas is the fact that three of them have K:Rb ratios about 2× greater than the other 17; this is shown in Figure 8. These three samples, together with two others, were taken for rare earth, K:Rb, and $^{87}Sr:^{86}Sr$ analyses by isotope dilution in the hope that these three would be abnormal in other respects. To date, however, the high K:Rb ratios are the only

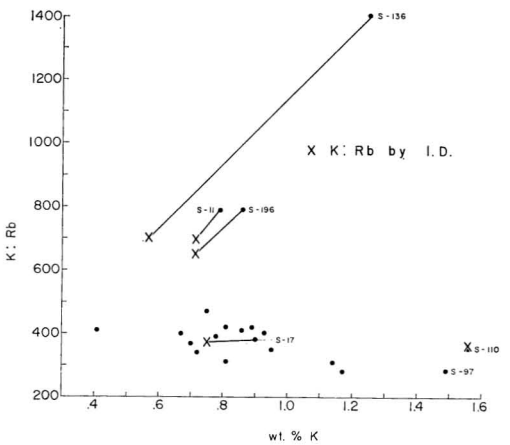

FIG. 8. K:Rb ratios vs. K contents for the 20 samples. The isotope dilution mass spectrometric data are shown by crosses.

unusual features of these three lavas. The isotope dilution K values are consistently lower than the K values given in Stice (1968), but the discrepancy for S-136 is huge (see Table 2).

These K:Rb data (excepting S-11, S-196, and S-136) show a decrease in K:Rb from about 400 to about 300, with K concentration increasing from about 0.7 to about 1.2 percent by weight. This is similar to what Shaw (1968) found for a much larger group of igneous rocks.

Zirconium and Titanium Dioxide Concentrations

It has been shown (Hubbard, 1969) that Zr and TiO_2 are covarient in Hawaiian and oceanic ridge basalts. This is also true of these Samoan basalts (Fig. 9). However, the accompanying inverse correlation found by Hubbard for TiO_2 and Al_2O_3 is not found in these Samoan basalts. Instead TiO_2 and Al_2O_3 concentrations increase together as seen in Figure 10, and thus must be influenced by different factors. It was shown earlier that variations in Sr and Al_2O_3 concentrations are due to movement of olivine crystals. This effect has also caused much of the observed range in TiO_2 and Zr concentrations. Two outstanding features of the TiO_2 vs. Al_2O_3 plot are: (1) the basalts from Ofu and Olosega have higher TiO_2 concentrations than other Manu'a basalts for the same Al_2O_3 concentrations; and (2) the hawaiites and the feldspar-

TABLE 2

TRACE ELEMENT DATA AND K:Rb RATIOS

Sample	Parts per Million				
	Zr	Sr	Ni	Rb	K:Rb*
S-11	318	505	88	10	790
S-17	263	440	420	24	380
S-44	270	490	130	21	420
S-62	238	425	250	20	390
S-89	175	390	755	19	370
S-93	258	510	125	27	350
S-97	510	680	55	52	285
S-107	288	515	64	21	410
S-110	413	960	< 5	44	355
S-117	232	400	315	16	470
S-129	295	430	360	19	420
S-131	188	330	640	10	410
S-136	440	430	165	9	1400
S-156	282	465	505	26	310
S-196	360	545	90	11	790
S-207	195	350	1000	21	340
S-210	372	690	5	37	310
S-221*a*	215	380	500	17	400
S-226	300	500	110	23	400
S-229	330	590	68	42	280

* K data are from Stice (1968).

phyric basalt show decreasing TiO_2 with increasing Al_2O_3, probably due to the crystallization and settling of either ilmenite or TiO_2-rich clinopyroxene. By contrast, Tongan lavas have less than 60 ppm Zr and correspondingly low TiO_2 values (Hubbard and Gast, in prep.).

Hubbard (1969) has described the use of Al_2O_3 and TiO_2 data to infer relative and approximate depths of magma generation. In a later section it will be shown that the original magma for these volcanoes had about 12.3 percent Al_2O_3 and 3.5 to 5.0 percent TiO_2. This Al_2O_3 value is lower than that probable for initial Kilauea liquids (about 13.3 percent) and the TiO_2 values are higher, indicating a source depth greater than the 40 to 60 km inferred for Kilauea Volcano (Eaton and Murata, 1960).

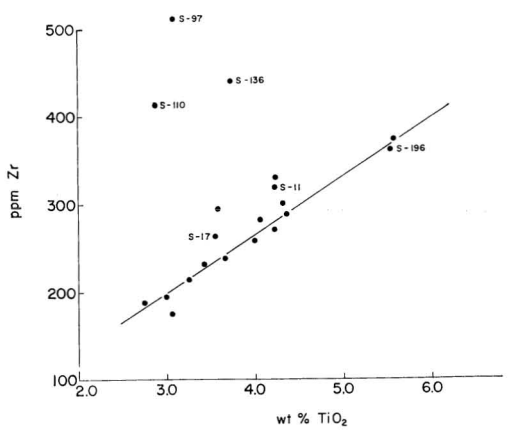

FIG. 9. TiO_2 vs. Zr for the 20 samples.

In Figure 9 one can see that S-110, S-97, and S-136 plot outside the scatter of the other data points. It is expected that S-110 and S-97 will also do so, but not S-136. Recall that the K concentration for S-136 is grossly uncertain; perhaps the two discrepancies have a common cause. Otherwise all seems normal with S-136.

Chemical Composition of the Original Magma

The term "original magma" will be used to indicate the liquid phase after segregation from the mantle material with which it last equilibrated and before the onset of crystallization at lower pressures.

Since the chemical composition of these lavas

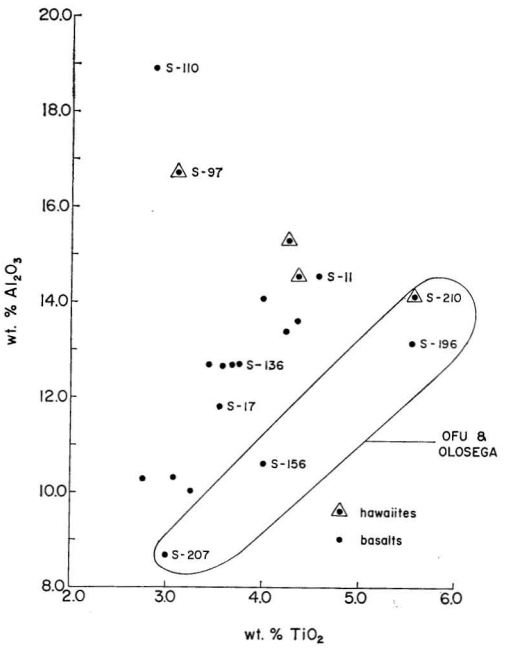

FIG. 10. TiO_2 vs. Al_2O_3 for the 20 samples.

is primarily controlled by the movement of olivine, it is possible to derive the chemical composition of the probable original magma. It was shown in the discussion of Ni and MgO concentrations that the original magma may have had as much as 373 ppm Ni and 10.4 percent MgO. Starting with this MgO value, the concentrations of other elements can be read off the graphs or estimated by scanning the chemical analyses. Column 1 in Table 3 shows the chemical composition of the probable original magma. Column 2 shows the approximate composition of the magma at the intersection of the olivine-plagioclase control lines of Figure 5. Note that slightly different TiO_2, Zr, and SiO_2 values are given for Ofu-Olosega.

TABLE 3

SOME DERIVED MAGMA COMPOSITIONS

	PROBABLE ORIGINAL MAGMA	APPROXIMATE COMPOSITION OF MAGMA AT INTERSECTION OF OLIVINE-PLAGIOCLASE CONTROL LINES (Fig. 5)
SiO_2*	45.5 (44.0)**	46.5 (44.4)
TiO_2	3.5 (5.0)	4.2 (5.8)
Al_2O_3	12.3%	13.7%
FeO (total Fe)*	12.5%	11.5%
CaO*	10.8%	10.0%
MgO	10.5%	7.1%
Na_2O*	2.5%	2.6%
K_2O*	0.8%	0.9%
P_2O_5	0.51%	0.58%
Sum	98.9%	97.1%
Ni	373 ppm	140 ppm
Sr	468 ppm	530 ppm
Zr	230 (330) ppm	260 (385) ppm
Rb*	17 ppm	19 ppm

* The values for these oxides and elements are less certain than for the others.

** Values in parentheses are for Ofu-Olosega, assuming equal Al_2O_3 concentrations.

CONCLUSIONS

The Manu'a lavas have strontium concentrations and rare-earth-element abundance patterns that are typical of oceanic island alkalic basalts (Hubbard, 1969). According to the partial melting model of Gast (1968), oceanic island alkalic basalts are the result of 3 to 7 percent of partial melting. The Al_2O_3 and TiO_2 data suggest that the depth of segregation is in excess of 40 to 60 km (Hubbard, 1969). The frequently high olivine content of these lavas indicates that they were erupted at rates that were often as great as any observed in eruptions of Kilauea Volcano, Hawaii (Murata and Richter, 1966). The low pressure differentiation of the Manu'a magma was characterized by the crystallization and movement of olivine and the differentiation of the original magma into picrites on one extreme and into Al_2O_3-rich residual liquids (hawaiites) on the other extreme.

The genesis of these Samoan basalts from the Manu'a Islands and the factors thought to have controlled that genesis are typical of those inferred for oceanic island alkalic basalts. This is strong evidence that the upper mantle in the vicinity of the Tonga-Kermadec Trench is typical for oceanic areas. Thus the geochemical and volcanological features of the Tonga lavas are probably due to their different tectonic situation.

ACKNOWLEDGMENTS

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